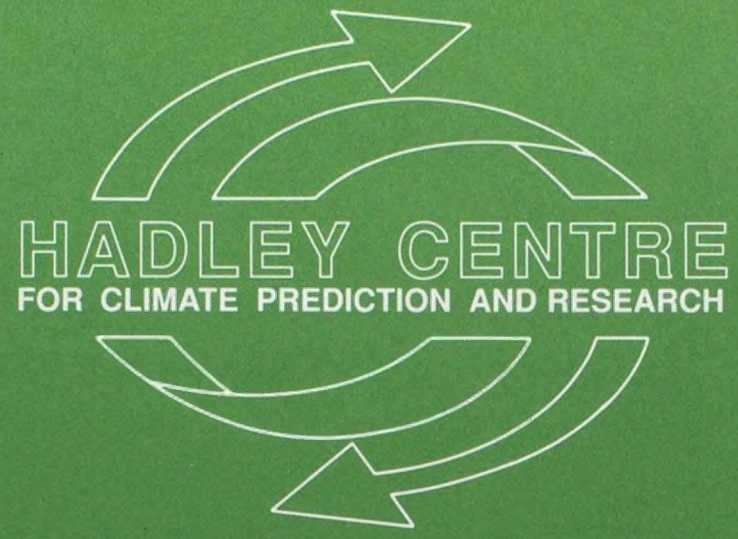

Simulation of global mean temperature using a box-diffusion climate model

by

P R Rowntree

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SIMULATION OF GLOBAL MEAN TEMPERATURE USING A BOX-DIFFUSION CLIMATE MODEL

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ABSTRACT

Prediction of climate requires understanding of past climate change. Radiative forcing of the climate by greenhouse gases alone cannot generate the observed twentieth century temperature record. It is shown here that, by using reasonable estimates of the contributions by greenhouse gases and anthropogenic and volcanic aerosols, a simple box-diffusion climate model simulates many features of the observed record without including either variations in solar intensity or internal variability. A good fit requires a climate sensitivity on doubling CO_2 of around 4K, within the range of estimates from general circulation models, but above the IPCC "best estimate" of 2.5K. Predictions to the mid-21st century show substantial sensitivity to the assumptions concerning future aerosol emissions.

INTRODUCTION

Variations in climate can be attributed to natural and anthropogenic perturbations of the radiative forcing and to internal variability (Manabe et al., 1991). Predictions of global and regional climate change require the use of three-dimensional coupled models of the circulation of the atmosphere and ocean and of the land surface. However, simpler climate models (HOffert et al., 1980; Wigley and Raper, 1992), which use a single multi-layered box to represent the global atmosphere and ocean can be used to simulate the global mean temperature record. These use estimates of the climate sensitivity - the equilibrium sensitivity of global mean temperature to a given radiative perturbation - which can be derived from the results of the three-dimensional models. Recent studies of climate change using such models have been able to fit the global mean temperature record only by allowing hypothetical contributions, as proposed by Friis-Christensen and Lassen (1991), from perturbations of the solar forcing (Schlesinger and Ramankutty, 1992; Kelly and Wigley, 1992) or inferring a significant role for internal variations of the climate system (Schlesinger and Ramankutty, 1994). In this paper we report experiments in which only forcing by greenhouse gases and tropospheric and stratospheric sulphate aerosols are included and obtain a fit to the observed record which requires only a relatively minor role for variations in solar intensity and internal climate variability.

MODEL

The model used here (Rowntree, 1994) takes advantage of the rapid coupling between the troposphere and the near-surface layers of ocean and land to combine these parts of the climate system as in Wigley (1985). The relatively rapid interchange of

heat between the top "troposphere/surface" layers for ocean and land is represented by a simple mixing calibrated using observations of the seasonal cycle. The slow transfer of heat between these regions and deeper layers of the ocean and land are represented by vertical diffusive processes. The globally averaged upwelling compensating a high-latitude bottom water formation term, included in some box-diffusion models, is omitted. Comparisons by Bretherton et al. (1990) suggest that the results from experiments omitting these processes lie between those from experiments including it which use different assumptions concerning the temperature of the downwelling water; its omission reduces by two the number of empirical constants in the model. The model's layer boundaries are at 30, 125, 375, 875, 1875 and 3875m in the ocean, and at 1, 3, 7, 15, 31 and 63m depth below the land surface. The vertical diffusivity is taken as 10^{-4} m^2s^{-1} for the ocean (Hansen et al., 1984) and $2.4*10^{-7}$ m^2s^{-1} for land.

The radiative heat flux H(1) at the tropopause is formulated as in the feedback studies of Schlesinger (1986):

$$H(1) = \Delta Q - \Delta T(1) * (1-f) / G_0 \qquad (1)$$

where $G_0=0.3K(W/m^2)^{-1}$ is the sensitivity of surface temperature to H(1) in the absence of feedbacks other than the negative one due to temperature. $\Delta T(1)$ is the departure of the top layer temperature from the initial equilibrium state and ΔQ is the imposed radiative perturbation. The factor f represents the climate feedback parameter (Schlesinger, 1986); the equilibrium response to ΔQ is $\{G_0 \Delta Q / (1-f)\}$. Thus for a doubled CO_2 perturbation ΔQ of $4.3Wm^{-2}$ at the tropopause the response in the absence of other feedbacks is 1.29K. The response of 1.5K to 4.5K proposed by the IPCC (Houghton et al., 1990) corresponds to a range of f from 0.14 to 0.713.

RADIATIVE FORCING

Greenhouse gases

The model includes forcing due to greenhouse gases (carbon dioxide, methane and nitrous oxide) and tropospheric and stratospheric aerosols. The histories of the concentrations of greenhouse gases from 1750 to 1990 were based on estimates from ice core and surface-based data (Boden et al., 1991; Etheridge et al., 1992; Khalil and Rasmussen, 1992). CFCs and their depletion of stratospheric ozone were neglected as these effects may roughly cancel (Isaksen et al., 1992). Variations in solar emissions have been omitted because of lack of knowledge (Isaksen et al., 1992).

Tropospheric aerosols

Tropospheric aerosols were represented by sulphates. Because of their short lifetime in the troposphere, the sulphate concentration was assumed to be proportional to the sulphur emission rates (ES) (Dignon and Hameed, 1989; Hameed and Dignon, 1992). Tropospheric aerosols deplete solar radiation both directly - clear sky scattering - and indirectly -

enhancement of cloud reflectivity. Estimates of the recent tropopause radiative forcing for sulphur emissions of 70 MtS/year range between about 0.3 and 1 W/m^2 for both direct (Charlson et al., 1992; Kiehl and Briegleb, 1993) and indirect effect (Charlson et al., 1992; Kaufman and Chou, 1993; Jones et al., 1994). The dependence on emissions is expected to be roughly linear for the direct effect. For the indirect effect, the increase is less than linear (Kaufman and Chou, 1993; Martin et al., 1994); analysis of a series of experiments by A. Jones (personal communication) with different aerosol loadings led to use here of a proportionality to $((1+0.1*ES)^{0.21} - 1)$ similar to that of Kaufmann and Chou (1993).

Stratospheric aerosols

Stratospheric aerosols originating from volcanic eruptions also perturb the radiative forcing of climate (e.g. Russell et al., 1993). Here, year-to-year variations in volcanic aerosol optical depth have been based on the Lamb Dust Veil Index (Lamb, 1970) up to 1883, and subsequently on estimates of aerosol optical depth compiled by Sato et al. (1993), linked using data from ice cores (Hammer et al., 1980; Moore et al., 1991) as described in Appendix B. Conversion of the optical depth data to tropopause radiative forcing was based on calculations (Russell et al., 1993) following the eruption of Pinatubo, showing that above Mauna Loa observatory an aerosol optical depth of 0.1 gave a reduction in downward flux at the tropopause of about 2.5 W/m^2. This was reduced by 0.2 W/m^2 (W J Ingram, personal communication - based on calculations with the Hadley Centre GCM radiation scheme) to allow for the typical stratospheric warming of about 1K observed following recent eruptions (Angell et al., 1993), giving 0.23 W/m^2 for a 0.01 increase in aerosol optical depth. Global generalisation of this local result is justified to first order because increased solar zenith angles lessen the vertical flux but enhance the fractional depletion.

The decadally-averaged volcanic radiative forcing, shown by the variations of the total heating relative to the combined effect of greenhouse gases and tropospheric aerosols (Fig. 2), exhibits variations of from +0.3 to -0.5 W/m^2 relative to its 1750-1990 mean. It should be remembered that there are no radiative observations to support these estimates of the volcanic effects before 1883 and that the icecore data may have significant errors (e.g. Hammer et al., 1980).

Evolution of heating rates

Figure 1 shows the estimated contributions to heating rates. These indicate that till 1900 biospheric sources dominated the forcing by CO_2, with fossil fuel CO_2 dominant only after 1950. Methane provided a substantial contribution throughout. The total forcing due to greenhouse gases (Fig. 2) evolved fairly smoothly, with the increase generally accelerating except during periods of relatively small increases in fossil fuel burning (Table 1). When tropospheric aerosol forcing is included as in Fig. 1, total globally averaged radiative forcing is much

reduced (Fig. 2); its rate of increase declines towards 1914 as aerosol cooling balances warming by biospheric CO_2 and methane but is then revived by slowing of the fossil fuel increase. The forcing declines again after 1945 relative to that without aerosols, accelerating again in the 1970s. The uneven evolution of aerosol cooling is due to its effectively instantaneous response to variations in fossil fuel burning (see Appendix C).

Figure 3 shows that, during periods of relatively rapid increase (Table 1), as before 1914 and after 1946, the rapid aerosol response outweighs the slow increase in the radiative perturbation due to slow accumulation of CO_2. The CO_2 increase became dominant before the slowing of fossil fuel burning in 1973 because of conversion from sulphur-rich coal to oil and natural gas, and because of increasing constraints on emissions from coal burning (Dignon and Hameed, 1989); by 1980 the estimates from Dignon and Hameed (1989) and Hameed and Dignon (1992) indicate that these reduced the ratio of sulphate emissions to CO_2 emissions by about 31% and 18% respectively.

EXPERIMENTS AND RESULTS

Effects of greenhouse gases and tropspheric aerosols

Most of the experiments reported here have been run using a feedback factor f=0.7. This gives a climate sensitivity of 4.3K for doubling of CO_2, and, when tropospheric aerosols are included, a total modelled warming over the period from the late 19th century to 1990 which is of similar order to that observed (Boden et al., 1991). However, in the absence of aerosols, a much lower climate sensitivity f=0.2 (1.6K for doubling CO_2) is required to give a realistic warming since 1900; with this value of f the warming since 1750 reaches 0.15K about 1900 and 0.6K by 1990; these figures are doubled with f=0.7. The heating (Fig. 2) increases at an accelerating rate and the temperature response (Fig. 4(a)) is correspondingly simple and different from the observed. When tropospheric aerosols are included, the temperature changes (Fig. 4(a)) show similar variations to those seen in the forcing and more similarity to observations.

Effects of volcanic aerosols

Including volcanic aerosols considerably modifies the temperature variations. The varying frequency of eruptions introduces (Fig 4(b)):

(a) lack of warming from 1860s to 1910s (realistic)
(b) cooling around 1880s (not supported by observations)
(c) strong warming from 1910s to 1940s (as observed)
(d) lack of warming 1940s to 1970s (as observed)
(e) slower warming from 1970s to 1980s.

Comparison with the observed temperature record

Although the model can represent many features of the observed record of global mean temperature since the late 19th century, the absence of observed inter-decadal temperature variations from the 1850s to 1920 is not simulated. Possible explanations

for this include:
(a) occurrence of internal variations which opposed the volcanic forcing; the interdecadal variations are within the range of $\pm$0.1K from global models (Manabe et al., 1991);
(b) poor knowledge of the volcanic forcing; there are no radiation data to support the record before 1883 and only one station until late 1892 (Kimball, 1924; Sato et al., 1993);
(c) incorrect estimation of global mean temperatures at a time with poorer observational coverage than recently.
(d) Omissions of the uncertain variations in solar luminosity.
The rise of temperature from early to mid-19th century is consistent with the evidence in the glacier record (Folland et al., 1990).

The predicted decadal mean temperatures were compared statistically with the observed data (Boden et al., 1991) for the period 1900-90 when the observed coverage exceeds 50-60% compared with 20-30% in the 1860s (Folland et al., 1992) and there is more confidence in the volcanic record. The regression-derived linear trend was removed so that the analysis isolated the skill of the predictions of departures from this trend. Table 2 shows that the correlation is near zero when no aerosols are included, while inclusion of tropospheric and volcanic aerosols gives a correlation of 0.71 (significant at the 5% level). The similar calculation for 1860-1990 gives a correlation of 0.65 (significant at the 2% level).

The greater variability in the observed data is consistent with the absence of natural variability in the model. A previous estimate with such a model giving a similarly good fit (e.g. Kelly and Wigley, 1992) included the variations in solar intensity related to solar cycle length (Friis-Christensen and Lassen, 1991) which they scaled so as to optimise the fit; however, the suggested solar mechanism appears to lack a firm physical basis (Kelly and Wigley, 1992; Folland et al., 1992). Apart from the magnitude of the tropospheric sulphate forcing (discussed below) the only parameter varied in the present experiments was the climate sensitivity (f). However, the ocean vertical diffusivity (κ) is also an adjustable parameter; Experiment 3 (Table 2) was rerun with κ increased from 10^{-4} to $2.25*10^{-4}m^2s^{-1}$ as suggested by Schlesinger (1986). The correlation was reduced by less than 0.01 indicating small sensitivity. The smaller standard deviation and trend suggest that a larger climate sensitivity could give a better result with this κ.

Otherwise the main uncertainty attaches to the magnitude of the forcing by tropospheric sulphates. Eliminating the indirect effect of aerosols, while keeping the same total response to 1990 emissions, leads to a slightly earlier warming because the sulphate cooling is reduced until 1990. The maximum difference of about 0.06K (Fig 4(c)) occurs in the mid-20th century. Increasing the tropospheric aerosol effect to $1.5W/m^2$ for 70MtS/year emissions, nearer the upper bound of the direct and indirect estimates, gives much less warming for the same

feedback factor (Fig. 4(c)). There is a clear need for more evidence on the magnitude of the tropospheric sulphate forcing.

Predictions of future temperature changes

Three scenarios of future changes in tropospheric aerosols have been considered:
(i) aerosol emissions as a constant fraction of fossil fuel burning from 1990 attributing the whole effect to direct effects and assuming 1 W/m^2 cooling at 1990;
(ii) as (i) but an equal role at 1990 for direct and indirect;
(iii) aerosol emissions constant at 1990 levels.

Scenario (ii) could correspond to unconstrained sulphate emissions with coal as the main fuel, at least in the countries with the largest increases in burning of fossil fuels, while scenario (i) is included mainly to demonstrate further the importance of distinguishing between the direct and indirect effects. Scenario (iii) represents a return to the trend of the period from the 1940s to the 1970s when the ratio of sulphate emissions to CO_2 emissions from fossil fuel burning fell by 40-50% through conversion to oil and gas and controls on sulphate emissions; limitation of the harmful effects of emissions through acid rain are the most likely motivation for such constraints. The feedback factor is 0.7 for each of these experiments.

The assumptions concerning 1990-2050 increases in the greenhouse gases were designed to give a scenario similar to that of the "business-as-usual" scenario used by Houghton et al. (1990); emissions of CO_2 were assumed to increase at 1.5%/year respectively, while methane increased at 16.85ppb/year; for N_2O the emissions were assumed to retain the same proportionality to those of CO_2 as before 1990, and the sink allowed for a lifetime of 167 years. The volcanic forcing for each decade after 2000 was assumed to be equal to the longterm (1750-1990) mean, with a minor eruption early in each decade.

Figure 5 shows the consequent changes in decadally-averaged global mean temperature in the model. The warming is considerably accelerated when aerosol emissions are held constant, with a warming rate from the 1990s to the 2040s greater by over 40% at 0.31K/decade compared with 0.22K/decade. Figure 5 also shows the warming computed when aerosols are ignored, and the climate sensitivity is given a sufficiently low value (f=0.2) to give the same warming to 1990 as when tropospheric aerosols are included. The anticipated warming rate is then only 0.155K/decade.

SUMMARY

Inclusion of tropospheric and stratospheric sulphates and use of a positive feedback giving 4.3K equilibrium warming for CO_2 doubling has allowed probably as realistic a simulation of twentieth century temperatures as can be expected, in view of model estimates of internal climate variability (Manabe et al.,

1991). There appears to be no need for a contribution from the hypothetical variations in solar intensity suggested by Friis-Christensen and Lassen (1991); nor do regular internal variations of the climate system as suggested by Schlesinger and Ramankutty (1994) appear to be necessary to explain the climate record though this does not disprove their existence. The comparison with observed data suggests that the real climate sensitivity to doubled CO_2 is at the upper end of the range of 1.5 to 4.5K proposed in the 1990 IPCC Report (Houghton et al., 1990), although we cannot be sure that internal variations of the real climate are not either enhancing or limiting the observed warming to a substantial extent. Predictions to the mid-21st century show substantial sensitivity to future aerosol emissions.

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APPENDIX A

Heating due to greenhouse gases

The heating due to carbon dioxide (Q_C), methane (Q_M and nitrous oxide (Q_N) was calculated using

$$Q_C = (4.3/\ln 2)\ (\ln(CC)-\ln(CC)_0),$$
$$Q_M = 0.0424\ ((CM)^{0.5} - (CM)_0^{0.5}) \text{ and}$$
$$Q_N = 0.14\ ((CN)^{0.5} - (CN)_0^{0.5}),$$

where (CX) is the concentration of gas X and $(CC)_0$=278.4ppmV, $(CM)_0$=748ppbV and $(CN)_0$=280ppbV are initial (AD 1750) concentrations. These formulations (Wigley, 1985; Hansen et al., 1984) give very similar radiative perturbations to those used by IPCC (Shine et al., 1990); the omission of overlaps between methane and nitrous oxide makes little difference for nitrous oxide and is roughly compensated for methane by omitting indirect effects due to water vapour (Shine et al., 1990) and ozone (Isaksen et al., 1992).

The separation between fossil fuel and biospheric CO2 heating rates was obtained by use of a simple model fitted to impulse response curves (Schwartz, 1993), based on Maier-Reimer and Hasselmann (1987). Half the CO_2 input decays with a 130 year halflife, 29% is assumed to be lost effectively immediately to the ocean and terrestrial sinks and the remaining 21% is retained permanently. When used with the observed fossil fuel input and CO_2 history, this gives a biospheric input with similar shape to that obtained with a box-diffusion carbon cycle model (Siegenthaler and Oeschger, 1987), but a smaller maximum input of about 0.5 GtC/year.

APPENDIX B

Linking of volcanic emission histories

To link Lamb (1970)'s data set on volcanic aerosol to that of Sato et al. (1993), decadal means of volcanic sulphate depositions estimated for 1750-1970 from measurements on ice cores from Greenland and Antarctica (Hammer et al., 1980; Moore et al., 1991) were each normalised to give the same mean and standard deviations as the estimates (Sato et al., 1993) of aerosol optical depth for 1880-1970 and then averaged together to provide a global mean. The annual data were then adjusted by a constant ratio through each decade to give the decadal means estimated from the icecore data. These adjustments were mostly small after 1880, except for the decade 1910-1919 when the icecore data indicate higher values of conductivity than would be expected from the astronomical data, particularly during the second half of the decade; the decadal mean for this period was halved to limit the adjustments to the annual data. Also, minor adjustments were made to Lamb's data based on evidence from icecores concerning an eruption around 1809 and the Coseguina eruption of 1835 (Delmas et al., 1992).

APPENDIX C

Relative contributions of CO_2 and sulphate aerosols to effect of fossil fuel burning

If the total (sulphate and CO_2) forcing F(t) due to fossil fuel burning behaves like

$$F(t) = a \int_0^t ffi(t')dt' - b\ ffi(t) \qquad (2)$$

then balance between the sulphate (second term) and CO_2 forcing (first term) requires that the fossil fuel input (ffi(t)) be proportional to exp (at/b) (D L Roberts, personal communication). Thus with fossil fuel burning increasing exponentially, the sulphate cooling may balance or exceed the CO_2 heating, but when the increase of ffi(t) slows, the CO_2 heating can become more dominant. For the estimated fossil fuel burning (Table 1) and radiative forcing as defined above, (a/b) was large enough for sulphates to dominate till 1913 and for part of 1946-1973. The calculated history of the decadal increases in total anthropogenically forced heating (Figs. 2, 3) illustrates this; the CO_2 increase became dominant before 1973 because b decreased as constraints on sulphur emissions increased (Dignon and Hameed, 1989). Note that even with a balance between sulphates and CO_2 forcing in the global mean, climate would still be affected because of the spatial heterogeneity of the sulphate forcing.

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FIGURE CAPTIONS

Figure 1: Contributions to the radiative perturbation (W/m^2) at the tropopause relative to 1750 for methane (solid), nitrous oxide (dotted), fossil fuel CO_2 (dashed) and biospheric CO_2 (dash-dotted) and tropospheric sulphate aerosols (assumed to give 0.5 W/m^2 each for direct and indirect forcing by 70 MtS/year emissions) (dash-double-dotted).

Figure 2: As Figure 1, showing
(i) (solid) total for greenhouse gases;
(ii) (dashed) greenhouse gases plus tropospheric aerosols; and
(iii) (dotted) greenhouse gases plus tropospheric and volcanic aerosols.

Figure 3: Decadal increases in radiative forcing ($W\ m^{-2}$/decade) due to (i) (dashed) fossil fuel CO_2 and
(ii) (dotted) tropospheric sulphate aerosols giving 1 W/m^2 with 70 MtS/year emissions for direct effect and
(iii) (solid) as (ii) but with equal split between direct and indirect as in Figure 1.

Figure 4(a): Global mean warming as
(i) (solid) observed (from Jones and Wigley's data in Boden et al. (13), shifted relative to the 1854-1990 mean by +0.45K) and
(ii) (dash-dotted) from box-diffusion climate model for experiments with greenhouse gases but no aerosols using feedback parameter f of 0.2
(iii) (dashed) as (ii) but with f= 0.7
(iv) (dotted) with sulphate forcing as in Figure 1 (f=0.7).
Figure 4(b): Global mean warming as (i) (solid) observed and
(ii) (dotted) as modelled with box-diffusion climate model using forcing as Figure 2 (f=0.7).
Figure 4(c): Results of experiments with forcing
(i) (solid) due to greenhouse gases plus tropospheric and volcanic aerosols as in Figure 4(a),
(ii) (dashed) as (i) but no volcanos,
(iii) (dash-dotted) with 50% more tropospheric sulphates than in (i), and
(iv) (dotted) as (i) with 1 W/m^2 forcing for emissions of 70 MtS/year but all from direct aerosol effect.
Feedback parameter f=0.7 for all experiments.

Figure 5: Changes in global mean temperature predicted for the period 1960 to 2050, with different assumptions concerning the evolution of cooling by aerosols. These are with forcing
(i) (dotted) due to greenhouse gases plus tropospheric and volcanic aerosols as in Figure 4(a);
(ii) (solid) as (i) but with direct effect of tropospheric aerosols doubled and no indirect effect as in Fig 4(c) (iv);
(iii) (dashed) as (i) but with constant aerosol emissions from 1990;
(iv) (dash-dotted) greenhouse gases only (no tropospheric or stratospheric aerosols). Feedback parameter f is 0.2 for (iv). (Feedback parameter f is 0.7 for (i)-(iii)).

TABLES

Table 1: Fossil fuel emissions of CO_2 (ffi)

Period	Fossil fuel burning (GtC/year)
1750-1913:	ffi = 0.061 * exp(0.042*(y-1850)
1913-1929.5:	ffi = 0.908 * exp(0.0073*(y-1913))
1929.5-1946:	ffi = 1.025 * exp(0.021*(y-1929.5))
1946-1973:	ffi = 1.452 * exp(0.042*(y-1946))
1973-1990:	ffi = 4.638 * exp(0.015*(y-1973))

Table 2: Removed trend (dT/dt), coefficient of correlation with observed (r) and standard deviation (σ) for model experiments and observed (Jones and Wigley's data in (13)).

Expt No.	Tropospheric aerosols	Volcanic aerosols	f	κ (m^2s^{-1})	dT/dt (K/yr)	r	σ (K)
1	No	No	0.2	.0001	.0054	-.01	.037
2	Yes	No	0.7	.0001	.0043	.23	.043
3	Yes	Yes	0.7	.0001	.0051	.71	.038
4	Yes	Yes	0.7	.000225	.0044	.71	.030
Observed					.0059		.059

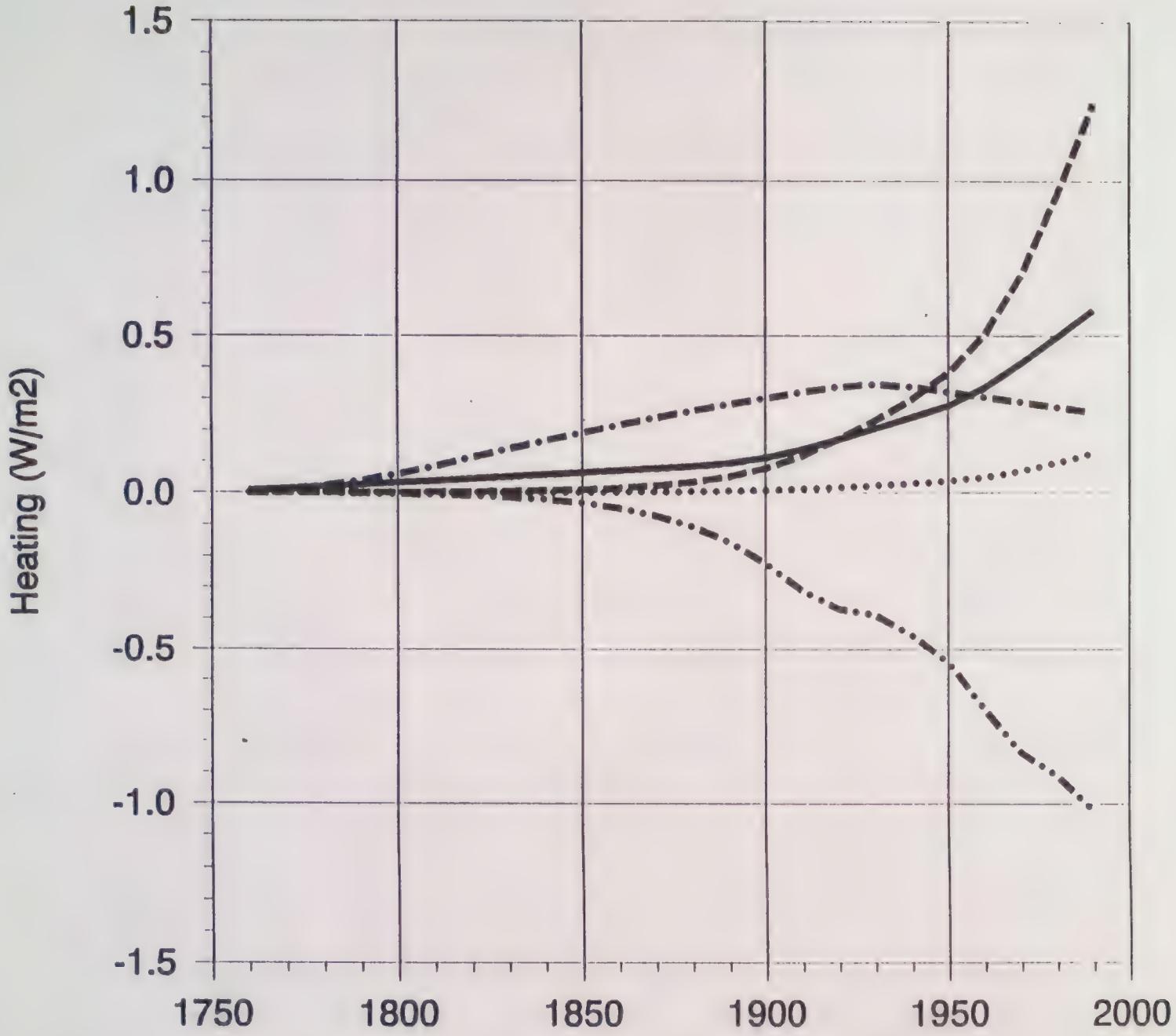

Figure 1

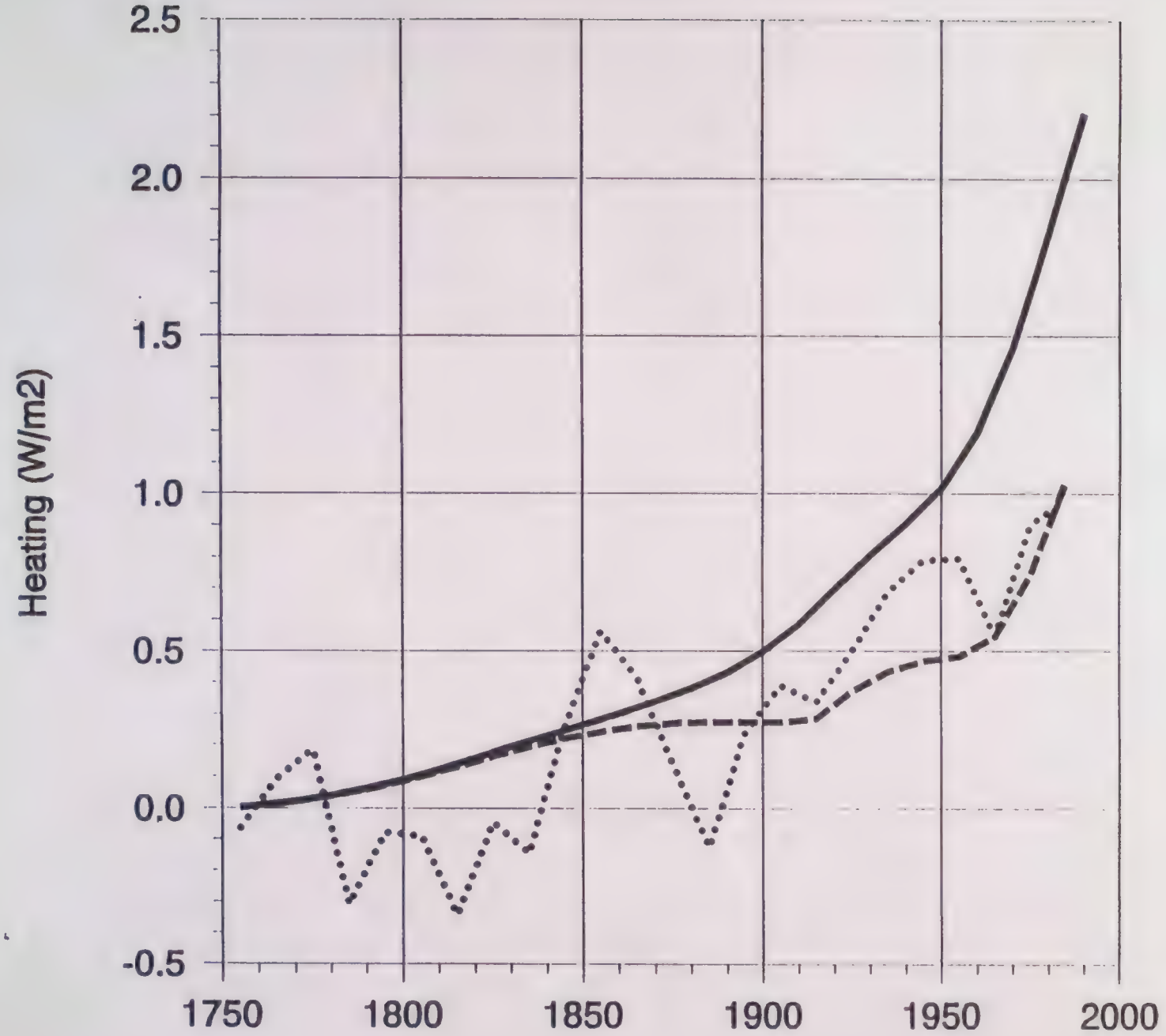

Figure 2

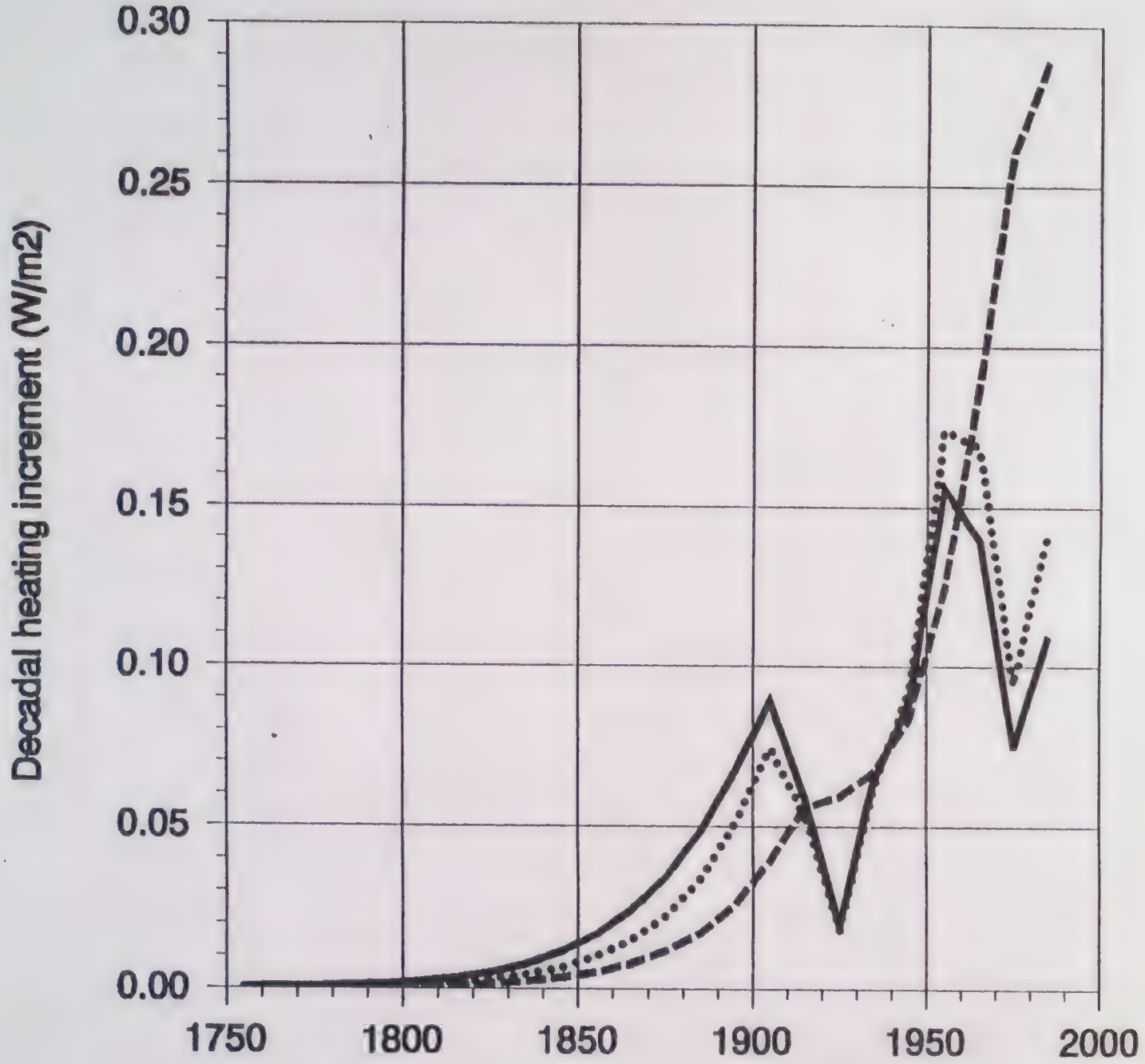
Decadal heating increment (W/m2)
0.30
0.25
0.20
0.15
0.10
0.05
0.00
1750
1800
1850
1900
1950
2000

Figure 3

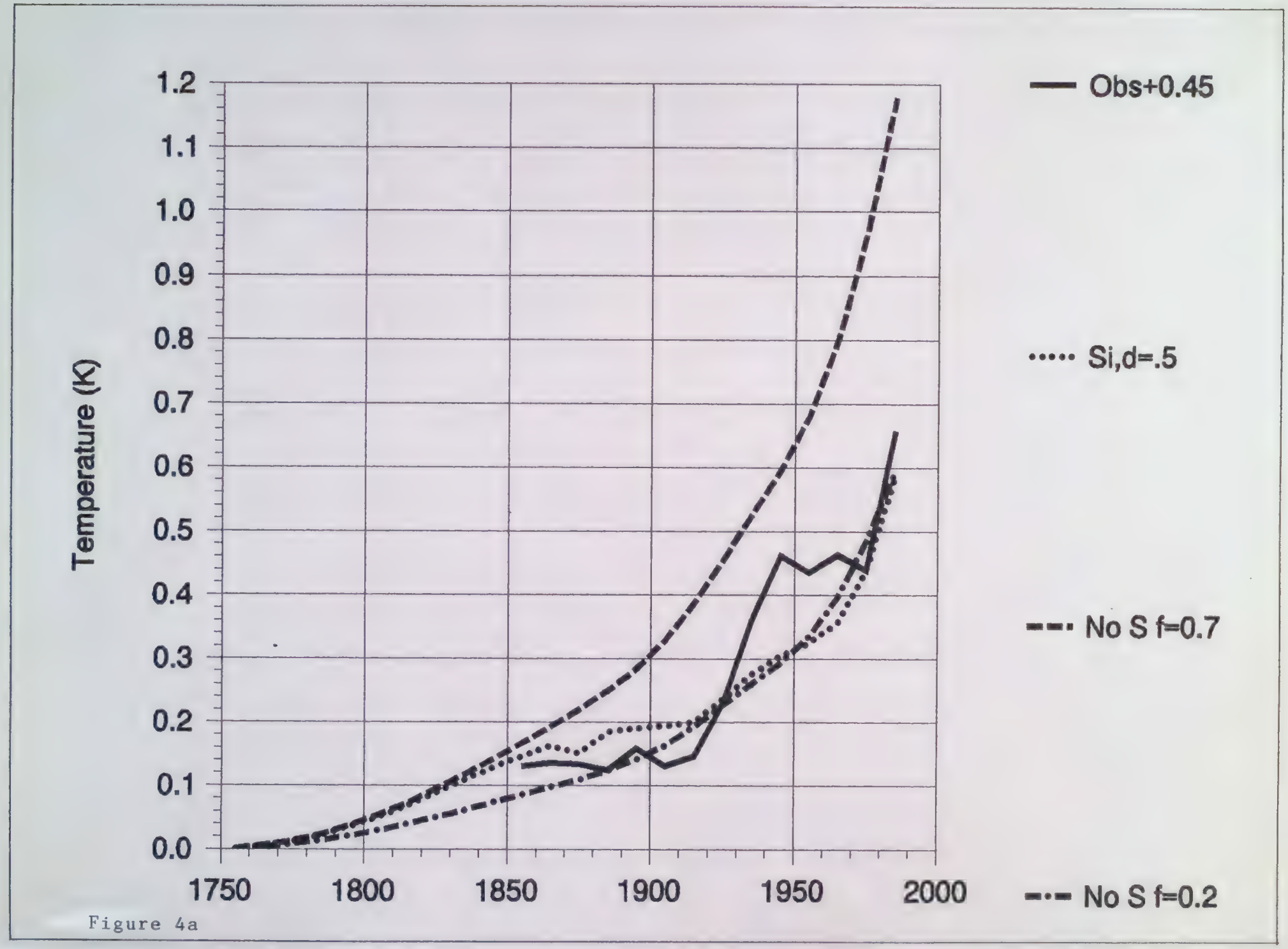
Temperature (K)
1.2
1.1
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
1750
1800
1850
1900
1950
2000
Obs+0.45
Si,d=.5
No S f=0.7
No S f=0.2

Figure 4a

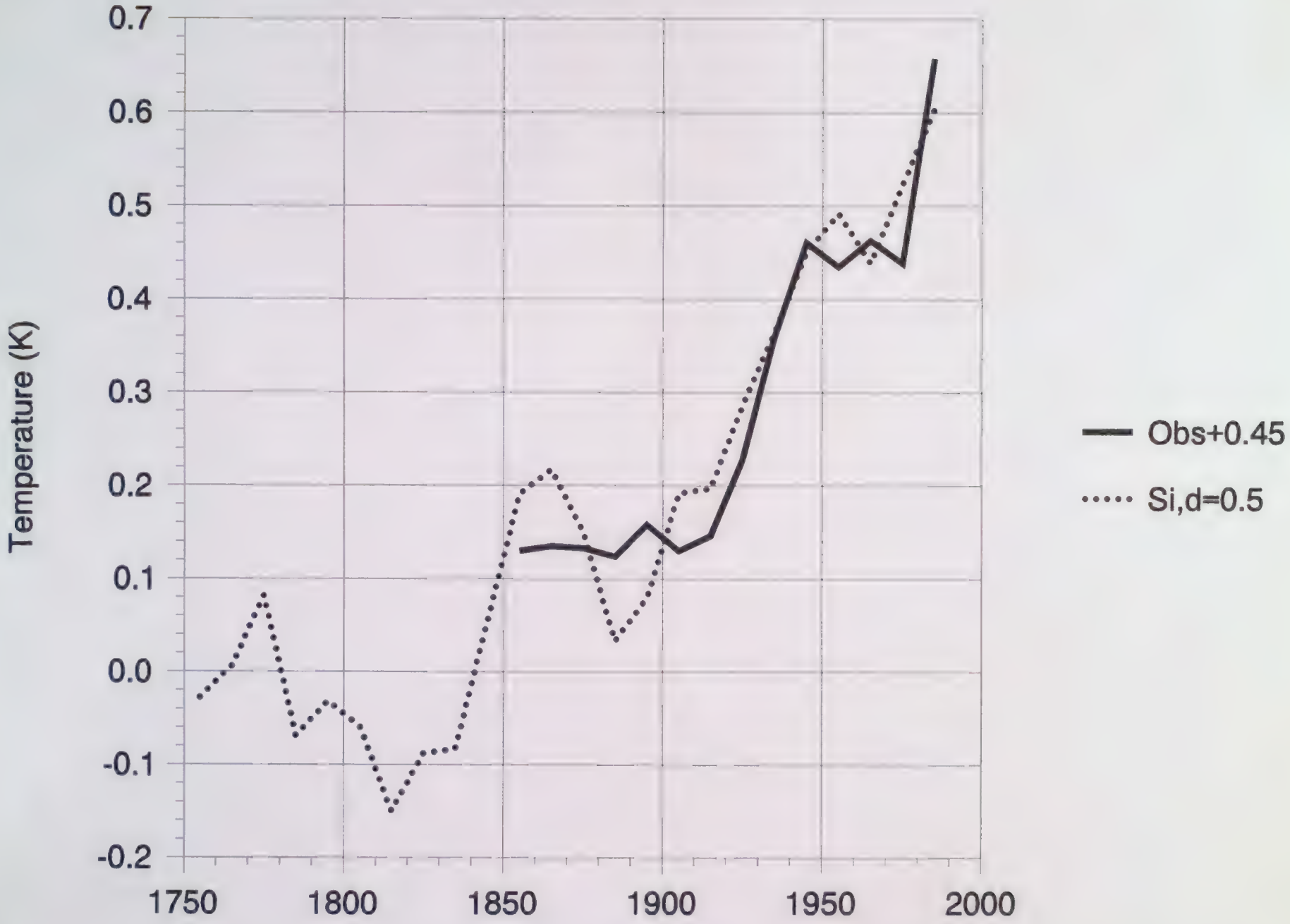
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
-0.1
-0.2
Temperature (K)
1750
1800
1850
1900
1950
2000
Obs+0.45
Si,d=0.5

Figure 4b

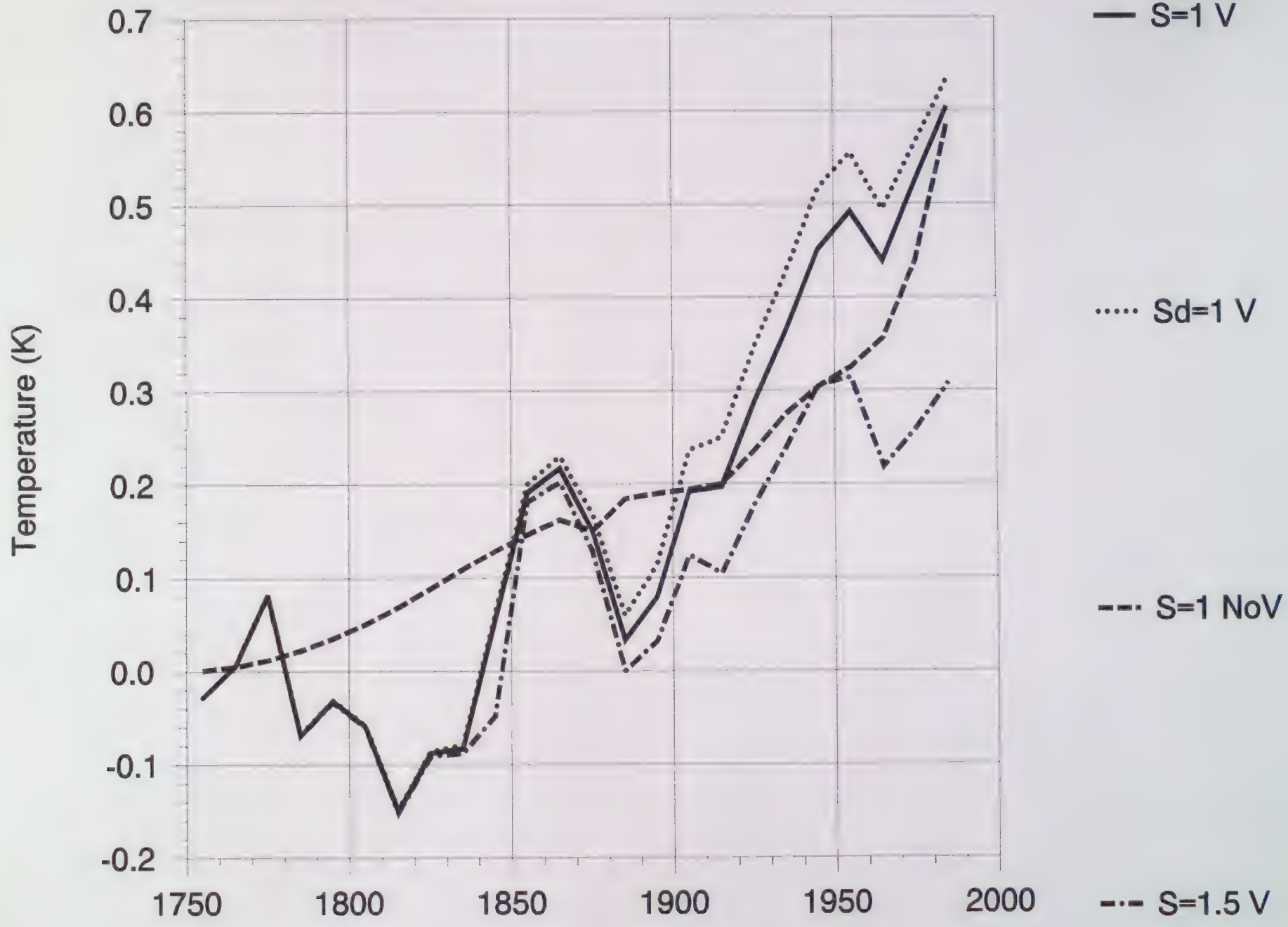
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
-0.1
-0.2
Temperature (K)
1750
1800
1850
1900
1950
2000
S=1 V
Sd=1 V
S=1 NoV
S=1.5 V

Figure 4c

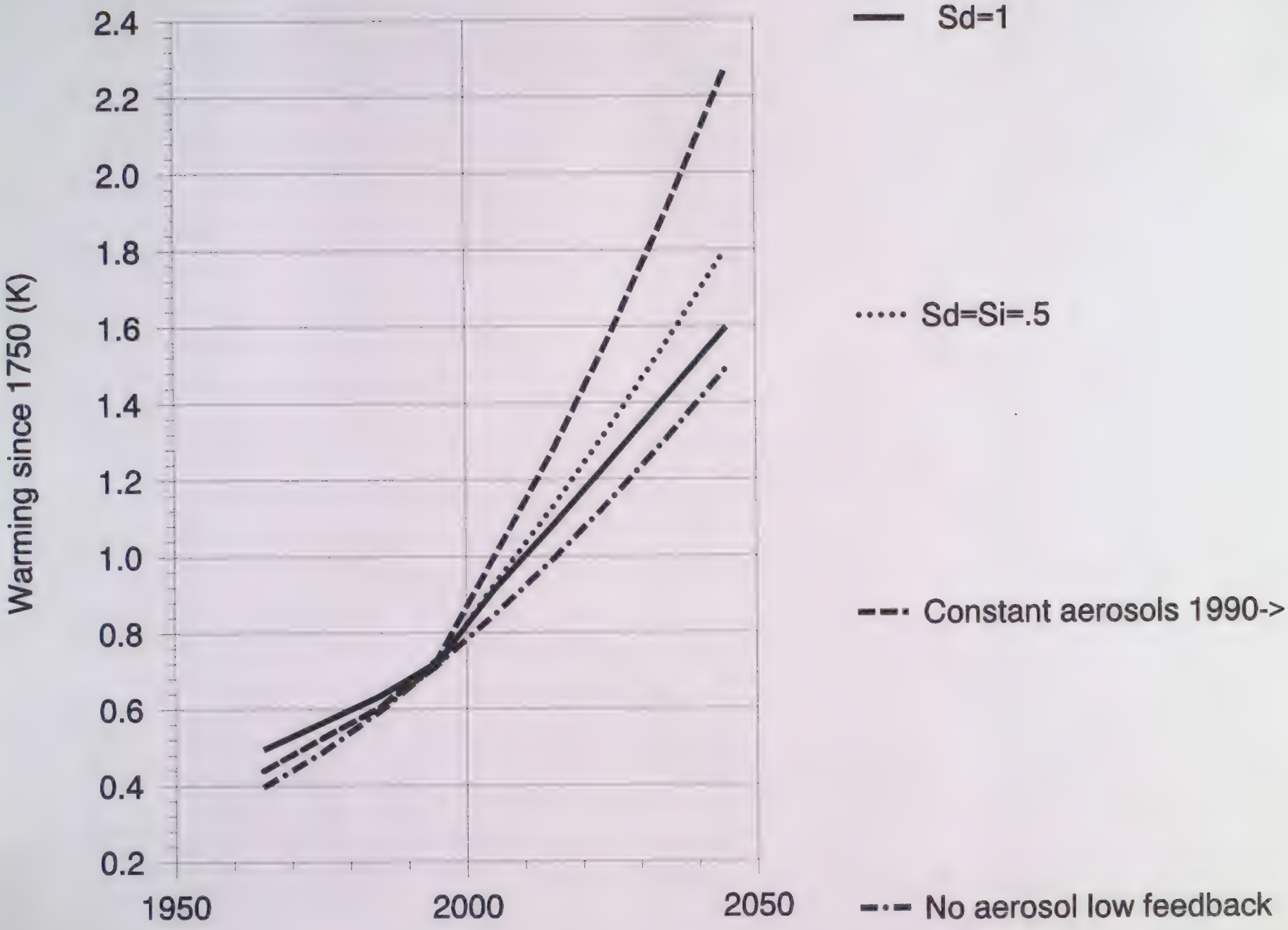
Warming since 1750 (K)
2.4
2.2
2.0
1.8
1.6
1.4
1.2
1.0
0.8
0.6
0.4
0.2
1950
2000
2050
Sd=1
Sd=Si=.5
Constant aerosols 1990->
No aerosol low feedback

Figure 5

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CRTN 50	Sep 1994	Simulation of global mean temperature using a box-diffusion climate model P R Rowntree